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LOCAL AND EXOTIC COMPONENTS OF
PRIMITIVE METEORITES, AND THEIR ORIGIN

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Abstract

Most of the material of chondrites was heavily reprocessed in the early solar system, and hence retains only a dim memory of its interstellar origin even in the least altered meteorites. Opaque matrix, the most primitive material, seems to have taken up Fe^{2+} and changed its mineralogy and texture. Chondrites and Ca,Al-rich inclusions have been further altered by melting, oxidation or reduction, loss of volatiles, etc. Nonetheless, small amounts of exotic components have survived, as indicated by isotopic anomalies. A dust component enriched in O^{16} is the most abundant and widespread. It survives in Ca,Al-rich inclusions as anomalous spinel grains, but is recognizable even in highly evolved meteorites and planets from variations in bulk oxygen isotopic composition. A few inclusions show small nucleosynthetic anomalies for many elements (Si, Ca, Ti, Cr, Sr, Ba, Nd, Sm), always coupled with mass fractionation of several of these elements, as well as O and Mg. Seven extinct radionuclides (Al^{26} , Ca^{41} , Mn^{53} , Pd^{107} , I^{129} , Sm^{146} , and Pu^{244}) have been recognized from their decay products, and provide clues to the chronology and nucleosynthetic sources of the early solar system.

Highly volatile elements, such as C, N, and the noble gases, show especially large and numerous isotopic anomalies. The noble-gas components include Ne-E (monoisotopic Ne^{22} from the β^+ decay of $2.6a \text{ Na}^{22}$), Xe-HL (enriched 2-fold in the light and heavy isotopes), and Xe-S (enriched in the even-numbered, middle isotopes 128, 130, and 132). They are located in carriers which themselves are anomalous, e.g. carbon enriched up to 2.4-fold in C^{13} or depleted by >30% in N^{15} , or spinel enriched in O^{16} and C^{13} . Other anomalies include nitrogen enriched 2-fold in N^{15} and hydrogen enriched 5-fold in D — probably relict interstellar molecules. A variety of astronomical sources seem to be required: novae, red giants, supernovae, and molecular clouds.

1. Introduction

All meteoritic matter originally came from outside the solar system and thus is exotic in the broadest sense of the term. However, much of this material was reprocessed in the early solar system (by vaporization, melting, mixing, isotopic exchange, chemical reactions, etc.), and became isotopically homogeneous. Such commonplace material is properly called "local," leaving the term "exotic" or "presolar" for material that still retains an anomalous isotopic signature. Exotic material sometimes occurs as discrete, potentially separable grains, but often has been assimilated by large amounts of local matter, and is recognizable only by its anomalous isotopic composition.

Qualitatively, a similar picture may be expected for comets. But as comets formed in more distant, cooler parts of the solar system, reprocessing may have been less intensive and pervasive. Thus the local material may be less processed and less dominant, with the exotic material more abundant and diverse. Nonetheless, primitive meteorites are a good frame of reference for interpretation of cometary data. I shall therefore first review the properties and classification of chondrites (§2) and then their local and exotic components (§3 and 4).

2. Properties and Classification of Chondrites

Chondrites are stony meteorites containing chondrules, mm-sized silicate spherules that appear to be frozen droplets of a melt. They consist largely of olivine $[(\text{Mg}, \text{Fe})_2\text{SiO}_4]$, orthopyroxene $[(\text{Mg}, \text{Fe})\text{SiO}_3]$, and plagioclase feldspar [solid solution of $\text{CaAl}_2\text{Si}_2\text{O}_8$ and $\text{NaAlSi}_3\text{O}_8$]. In the more primitive chondrites, glass occurs in place of crystalline feldspar.

Chondrules are embedded in a ground-mass or matrix. In the less primitive

chondrites, the matrix is somewhat finer-grained than the chondrules, but otherwise has the same mineralogy and composition. Millimeter-sized grains of metal (nickel-iron with 5-60% Ni) and troilite (FeS) are also present. In the more primitive chondrites, at least part of the matrix is very fine-grained (to $\sim 10^{-6}$ cm) and richer in Fe^{2+} than are the chondrules.

Five chondrite classes are recognized, differing in the proportions of oxidized to reduced iron. Enstatite (=E) chondrites are highly reduced, containing iron only as metal and FeS. Carbonaceous (=C) chondrites are highly oxidized, containing mainly Fe^{2+} , Fe^{3+} , and little or no Fe^0 . The middle ground is occupied by 3 classes of intermediate oxidation state and variable iron content: H, L, and LL chondrites. (The letters refer to total iron content: high, low, and low-low.) Collectively, they are often called ordinary (=O) chondrites.

Van Schmus and Wood (1967) have further subdivided each of these classes into "petrologic types," numbered from 1 to 6. These types were originally designed to reflect increasing chondrule-to-matrix intergrowth (probably due to thermal metamorphism in the meteorite parent bodies, at T up to 1100°C), but have turned out to correlate well with compositional trends, e.g., volatile content. Probably these types represent an accretion time sequence, the higher types having accreted earlier, and therefore being located in deeper and warmer locations in the parent body.

Only the most primitive chondrites, up to petrologic type 3, are of interest to us, as they are least affected by metamorphism and hence have best preserved their exotic components as well as a record of processes in the solar nebula. Regrettably, even these meteorites are not pristine. Types 1 and 2—found only among C-chondrites—have undergone hydrothermal alteration in

their parent bodies (DuFresne and Anders, 1962; Clayton and Mayeda, 1984). Type 3, in turn, has been slightly metamorphosed, at T up to ~300 - 400°C. Fortunately the least metamorphosed Type 3's can be recognized from their low thermoluminescence sensitivity and other criteria, and on this basis, Type 3's have been divided into 10 subclasses, 3.0 to 3.9 (Sears et al., 1980; Anders and Zadnik, 1985).

C3 chondrites contain a minor but very important textural component: mm-sized, often irregular inclusions of refractory, Ca, Al-rich but Si-poor minerals such as melilite, $\text{Ca}_2(\text{Al}_2, \text{MgSi})\text{SiO}_7$ or spinel, MgAl_2O_4 . They are known by the acronym CAI.

3. "Local" Components of Chondrites and Their Origins

To a first approximation, chondrites are condensates from the solar nebula, as first recognized by Wood (1958). Thus a convenient if oversimplified framework for interpreting chondrites is the equilibrium condensation sequence of a solar gas (Larimer, 1967; Grossman, 1972; Lewis, 1972; Grossman and Larimer, 1974). Fig. 1 shows this sequence. A more detailed version of the 300 - 1800K portion is given in Fig. 2. As chemical equilibrium is independent of the path, these diagrams apply equally well to isobaric heating or cooling and to isothermal compression.

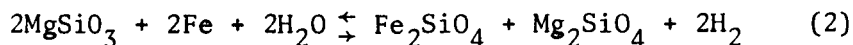
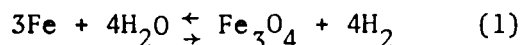
Fig. 1 helps rationalize the composition of planets and other small bodies in the solar system. The inner planets and the asteroids consist only of rock; comets and the satellites of the outer planets consist of rock and ice, whereas the outer planets consist of rock, ice, and gas.

This appealing picture is oversimplified in several respects. (1) The various chondrite classes differ somewhat in bulk composition (Fe/Si, Mg/Si, Ca/Si, etc.), and so at best only one class can be a total condensate. All others must have been fractionated, presumably by gain or loss of major

condensates (Larimer and Anders, 1970). (2) The nebula was not wholly gaseous; some presolar solids survived (Clayton et al., 1973, see also §4). (3) Chemical equilibrium may not have been reached, especially at lower temperatures. (4) The composition may have deviated from solar, owing to gas-dust fractionations, etc. Nonetheless, the condensation diagram is a good frame of reference, yielding detailed predictions that can be compared with observations. I shall review the principal components of chondrites against this framework.

3(a) Matrix

Only a few meteorites of petrologic types 3.0 to 3.4 have retained pristine matrix, unaffected by metamorphism or hydrothermal alteration (Wood, 1962; Huss et al., 1981; Nagahara, 1984). This opaque (or "Huss") matrix contains micron- to submicron-sized silicate grains, amorphous material, and submicron-sized grains of Fe-Ni, FeS, and Fe₃O₄. The Fe₃O₄ and the high Fe²⁺-content of the silicates (the olivine has up to 50 mol % Fe₂SiO₄), are consistent with equilibrium in a solar gas at ≤ 400 K (Grossman, 1972):



In a solar gas, reaction (1) goes to the right below 400K, whereas reaction (2) reaches 50 mol % Fe₂SiO₄ at 470 \pm 40K.

Was the matrix totally reprocessed in the solar nebula or has it retained some of its exotic features? At least some of the minor components of matrix must be exotic, since they are isotopically anomalous (§4(d)). However, we know embarrassingly little about the bulk of opaque matrix, as it has been recognized only recently and has not received much study—in contrast to the

more common but less primitive types of matrix (Ashworth, 1977; Hutchison et al., 1979). One feature that suggests major reprocessing in the solar nebula is the high Fe^{2+} and Fe_3O_4 content. Reactions (1) and (2) are heterogeneous, and are barely fast enough at the high temperatures (400 - 500K) and densities [$n(\text{H}_2) \approx 10^{14}/\text{cm}^3$] of the solar nebula (Larimer and Anders, 1967). They would be many orders of magnitude too slow at the conditions of a molecular cloud ($T < 50\text{K}$, $n \approx 10^6/\text{cm}^3$). The required combination of moderate T and large $n(\text{H}_2)$ is not available in any known astrophysical environment other than the solar nebula.*

Actually, reactions (1) and (2) could proceed at higher temperatures if the $\text{H}_2\text{O}/\text{H}_2$ ratio were greater than solar. The easiest way to raise this ratio would be by a dust/gas fractionation, since dust is a major carrier of oxygen (Larimer and Bartholomay, 1983; Wood, 1984). As dust would tend to settle toward the nebular midplane where accretion takes place, some such fractionation is expected. We shall return to this problem in §3(b).

Granted that Fe^{2+} and Fe_3O_4 contents of matrix were established in the solar nebula, it does not follow that all other features were, too. Very little work has been done on matrix; in particular, no systematic search has been made for surviving interstellar features. Several isotopically anomalous, hence presolar components have been found in chondrites

*There is a minor loophole in this argument. If the iron in interstellar grains was very finely dispersed (monolayer or individual atoms) at some stage of its history, then it may have become oxidized prior to its arrival in the solar system. However, such ultrafine material would react with the nebular gases during its reorganization to larger crystals, and so its final oxidation state would still be established in the nebula.

[C, H, N, and noble gases; §4(d)] and though none of them have been seen in situ — or otherwise conclusively linked to matrix — their presence shows that at least some heat-sensitive presolar components got into meteorites without serious damage. Thus it is quite conceivable that matrix of primitive meteorites is only lightly reprocessed interstellar dust (Wood, 1963, 1981; Anders, 1965; Cameron, 1973; D.D. Clayton, 1982; Huss, 1986). This is a testable proposition that should be checked experimentally.

3(b) Chondrules

Chondrules formed by flash melting of pre-existing solids, not by direct condensation from gas (Nagahara, 1981; Wood, 1984). The melting time was short (seconds to minutes), judging from the failure of three reactions to go to completion: loss of Na and other volatiles (Schmitt et al., 1965; Tsuchiyama et al., 1981), reduction of Fe^{2+} to Fe by nebular hydrogen, (Wood, 1984), and isotopic exchange with oxygen of the nebula (R.N. Clayton, 1981). The cooling time likewise was short (seconds to hours, depending on textural type), as shown by laboratory experiments (Tsuchiyama et al., 1980; Planner and Keil, 1982; Nagahara, 1983).

The short cooling times imply that the chondrule-forming region was small, allowing rapid heat loss by radiation. The existence of 5 major chondrite classes requires at least 5 separate domains, differing in oxidation state, bulk chemistry, and oxygen isotope composition.

Chondrules are complementary to other major components of the meteorite (metal, total matrix), being depleted in siderophiles and volatiles but enriched in refractories (Grossman and Wasson, 1983, 1985; Wood, 1984). Interestingly, the bulk chondrite always is close to solar composition,

which suggests that these components are cogenetic and accreted comprehensively. The parent material may have been the opaque matrix, perhaps after a preliminary stage of reprocessing (Rambaldi and Wasson, 1984). It is important to distinguish this primitive, opaque matrix from more fractionated and recrystallized types, which probably represent "failed chondrules": material that overshot or undershot the melting range (Wood, 1984).

Chondrule formation seems to have involved some recycling, as demonstrated by relict crystals (Nagahara, 1981; Rambaldi et al., 1983) and compound chondrules (Gooding and Keil, 1980). It was followed by accretion of rims of matrix-like material (Allen et al., 1980).

It seems that chondrule formation was a pervasive process that occurred throughout the inner solar system and reprocessed some 50 - 90% of the material. The inner planets are depleted in some of the same elements as the chondrites (e.g. alkalis, S, Se, Ga, Ge, etc.) and by Occam's Razor, it seems reasonable to invoke the same process (Morgan and Anders, 1980).

The mechanism of chondrule formation has not yet been conclusively identified. The complementarity of chondrules and matrix speaks against models that make chondrules and matrix in different places and mix them afterwards. The most promising model involves heating of interstellar matter as it falls into the solar nebula. Direct "infall heating" by aerodynamic drag (Wood, 1984, 1985) turns out to be quantitatively inadequate, but heating by viscous interaction with the nebula seems more promising (Wood, 1986). The chondrule-forming zone in this model is a thin, hot layer of the nebula, as implied by the short cooling times of chondrules. However, this model is very recent, and still has to pass quantitative tests.

The redox state (i.e., Fe^{2+} content of olivine) of primitive chondrules

is quite variable even within a single chondrule, ranging from less than 1 to 40 atompercent. At 1500 K in a solar gas, the equilibrium value would be only 0.001%, and so either the gas was non-solar or the parent material was highly oxidized and did not fully equilibrate with the ambient gas upon melting. Wood (1984), who favors the former alternative, has shown that Fe^{2+} contents of 0.5 to 30 atom percent could be attained if dust were enriched 10^2 to 10^4 -fold relative to gas, thus raising the $\text{H}_2\text{O}/\text{H}_2$ ratio in the system. However, it seems unlikely that chondrules fully equilibrated with the nebula, as they did not reach uniform Fe^{2+} contents, did not lose all their volatiles, and did not isotopically equilibrate their oxygen. More likely, chondrules formed from an Fe^{2+} -rich precursor (opaque matrix or a reprocessed derivative thereof), which was partially reduced and devolatilized on melting.

Even for the highly reduced E-chondrites, it appears that the precursor material was relatively oxidized (Rambaldi et al., 1984). Conditions must have become more reducing with time (Grossman et al., 1985), presumably due to a dust-gas fractionation in the nebula that raised the C/O ratio above the critical value of 0.8 (Larimer, 1975; Larimer and Bartholomay, 1979, 1983).

3(c) Ca, Al-Rich Inclusions ("CAI's")*

Chemically and mineralogically, CAI's look like an early condensate from a solar gas (Larimer and Anders, 1970; Marvin et al., 1970; Grossman, 1972).

*The literature on this topic is vast. For a more thorough treatment, see Grossman (1980), Wood (1981), Wark and Lovering (1982), Kornacki and Wood (1984), and Huss (1986).

They contain the suite of refractory minerals and trace elements expected to condense in the first ~6% of the "rock" fraction, between 1800 and 1400 K (Fig. 2). However, Fig. 2 applies equally well to evaporation and to condensation, and so some authors have argued that CAI's are evaporation residues (of presolar matter) rather than condensates (Kurat, 1970; D.D. Clayton, 1975; Wood, 1981).

It turns out that the truth lies in between, perhaps closer to the latter view. In the most conclusive test thus far, Niederer and Papanastassiou (1984) have shown that the isotopic fractionation patterns of Ca and Mg require a complex sequence of both processes. Of course, both are natural extensions of the chondrule-forming process. In some cases, this process overshot the mark, vaporizing part of the chondrule and leaving a residue enriched in refractories such as Ca, Al, and Sc (Osborn et al., 1974). As the vapor would have recondensed eventually in the (generally cool) nebula, the chondrule-forming process thus would yield condensates, in addition to evaporation residues and fully or partially melted chondrules.

Two main classes of inclusions are recognized on the basis of grain size. Coarse-grained inclusions are uniformly enriched in refractory elements to about 17x of CI-chondrite levels (Grossman et al., 1977), and hence may be interpreted as a total condensate of the first 6% of "rock" (Fig. 2) or as an equilibrium evaporation residue of the last 6%. The latter view seems to be closer to the truth, judging from the presence of isotopically anomalous spinel [§4(a)], which could not have condensed from or equilibrated with the nebula.

Fine-grained inclusions have highly fractionated abundance patterns, being enriched in refractories (but by variable factors) and in volatiles

such as alkalis and halogens (Grossman and Ganapathy, 1976). They contain not only the usual refractory minerals, but also feldspathoids such as nepheline (NaAlSiO_4) and sodalite ($3\text{NaAlSiO}_4 \cdot \text{NaCl}$). In terms of the condensation diagram (Fig. 2), they can be described as discontinuous condensates, which lost material condensing between 1800 and ~1600 K (light lanthanides and other highly refractory elements) and 1400 - 1200 K (metal and silicate), but collected elements condensing in other intervals, sometimes disproportionately so.

Some important if ill-understood clues to the formation conditions of coarse-grained CAI come from "Fremdlinge"* — 10 to 100 μm aggregates containing metal; alloys of Pt metals, Ga, Ge, Mo, Sn, and Re; Ca-phosphates; and sulfide or oxide minerals of other rare elements such as Sc, V, Zr, Nb, Mo, and W (ElGoresy et al., 1978; Armstrong et al., 1985). Some alloys can be rationalized in terms of condensation temperature, structure, or chemical reactivity; others, such as Os Ru Rh, cannot ($T_{\text{cond}} = 1910, 1650, \text{ and } 1470 \text{ K}$ at 10^{-3} atm ; structure hcp for Os, Ru but fcc for Rh).

It seems that these grains condensed under conditions highly conducive to nucleation and hence low supersaturation. Thus atoms could choose their preferred condensation sites and even a rare element such as Re was able to form its own phase rather than alloying with the previously condensed, structurally similar and more abundant Os. Yet owing to the rarity of these metals (solar $\text{Os}/\text{H} = 2 \times 10^{-11}$), it would take some 600 years for a $1 \mu\text{m}$ Os grain to form at $P = 10^{-4} \text{ atm}$, correspondingly less at higher P (Palme and Wlotzka, 1976). It is hard to see how the required high temperatures were sustained for so long a time, and how they were regulated to maintain low supersaturation. But the growth rate is proportional to $1/P$, and so

*German for "strangers," to connote a separate, possibly extrasolar, origin.

shorter times would suffice at higher pressures. Perhaps this is an argument for models postulating higher pressures (protoplanets; Cameron, 1978; denser gas regions surrounding planetesimals; Hayashi et al., 1985).

To complicate matters, the oxide, sulfide, and metal phases in Fremdlinge require a very wide range of oxygen fugacities and temperatures (Armstrong et al., 1985). Most of the trace elements in question are about as rare as Os, and so the same dilemma about growth time applies, but with the added complication of a wide spectrum of "microenvironments," differing in oxygen fugacity and gas composition. The problem can't be solved by relegating it to some presolar stage, as no known astronomical object offers the required combination of temperature, density, time, and composition. Supernovae, in particular, are too tenuous and short-lived for significant grain growth. Though abundant elements such as Mg or Al can yield micron-sized condensates in the few months available for condensation (Lattimer et al., 1978), rare elements such as Os cannot; an average Os atom will have only 10^{-12} collisions with another Os atom during the entire cooling interval from 1600 to 0 K.

By default, it seems necessary to invoke the solar nebula or the meteorite parent bodies as the site of the "microenvironments." If the alloys formed by condensation, then the required chemical diversity (especially the high oxygen fugacities) could be achieved either in the nebula itself or in the proto-atmosphere of growing asteroids or protoplanets (Cameron, 1978; Hayashi et al., 1985). In the nebula, high oxygen fugacities could be produced by gas-dust fractionation; in a protoatmosphere, by evaporation of accreting ice. In both settings, further variety would result from settling of dust through gas regions depleted in certain elements by prior condensation.

On the other hand, if the alloys formed by exsolution during metamorphism in the meteorite parent body (Armstrong et al., 1985), then the required changes in chemical environment could be caused by migration and escape of major volatiles, such as CO, H₂O, H₂S, etc. In any event, it seems that the "Fremdlinge" are not so "fremd," after all.

3(d) Carbon and Organic Matter

Primitive meteorites contain up to 4% C, with organic carbon dominant in C1's and C2's, but elemental carbon dominant in C3's (Table 1). Abundances of total C, H, and N also decline in this order, with a sharp drop beyond C3. (As C-chondrites of higher petrologic types are very rare, H4 to H6 chondrites have been substituted.) The condensation temperatures from the solar nebula have been estimated from various "cosmothermometers," such as isotopic fractionations or trace element abundances (Larimer, 1978).

The organic carbon consists mainly of an intractable, aromatic polymer similar to coal in its structural units (Hayatsu et al., 1977), and a variety of extractable organic compounds, including alkanes, alkenes, arenes, alcohols, carboxylic acids, amino acids, nitrogen heterocyclics (including the purine and pyrimidine bases of DNA and RNA), etc. (Hayatsu and Anders, 1981).

The alkanes consist mainly of the normal, straight-chain isomers, with lesser amounts of 4 or 5 slightly branched isomers (mono- and di-methyl). As there exist many thousands of possible isomers (e.g., 10,359 at C₁₆), the dominance of only 5 or 6 implies a highly selective process — presumably the Fischer-Tropsch synthesis, which involves catalytic hydrogenation of CO. Indeed, laboratory experiments show that all organic compounds reliably identified in meteorites can be produced from CO, H₂, and NH₃ in the presence of catalytically active meteoritic minerals, such as Fe₃O₄ and clays (Studier et al., 1968;

Hayatsu and Anders, 1981).

Fig. 3 illustrates the thermodynamics of this process in a solar gas. Let us consider isobaric cooling at 10^{-5} atm, although isothermal compression at ~400 K would give similar results.

At high temperatures carbon exists mainly in the form of CO. Near 600 K, hydrogenation to CH₄ should commence, and be 50% complete at 590 K. However, this reaction is very slow in the absence of catalysts, and as the dominant minerals at this temperature (Fe, Mg-silicates, FeNi coated with FeS) are very poor catalysts (Anders et al., 1973), most of the CO survives metastably.

At 520 K, disproportionation to C and CO₂ should set in. Again, good catalysts are lacking, and so much of the CO survives. Finally at 400 K, the anhydrous mineral assemblage dominant up to that point (Fig. 2) transforms to clays and Fe₃O₄, both of which are good catalysts for the hydrogenation of CO. However, at this low temperature, CH₄ — though still the most stable product — is no longer the only possible one. Many other organic compounds can form metastably, and actually do, being favored by the reaction mechanism.

The formation of organic compounds thus is triggered by the appearance of catalysts at 400 K. Indeed, meteorites rich in clay minerals (C1, C2) are rich in organic compounds, whereas meteorites poor in clay minerals (C3) have little or no organic matter, containing most of their carbon as elemental C (Table 1).

The aromatic polymer forms as a secondary product in this scheme, by aromatization of the primary aliphatic products. Such aromatization is known to occur on Fe₂O₃ (Galwey, 1972), and has also been seen in a Fischer-Tropsch synthesis extended over 6 months (Hayatsu et al., 1977).

4. Exotic Components of Chondrites

4(a) Oxygen-16

The most abundant and ubiquitous exotic component is O^{16} (Clayton et al., 1973; Clayton, 1981). Its presence can be demonstrated by a 3-isotope diagram, a variant of which is shown in Fig. 4. On this diagram, all samples related by mass fractionation lie along a line of slope 1/2 (reflecting the mass differences $O^{17}-O^{16}$ and $O^{18}-O^{16}$), whereas mixtures of independent components lie along "mixing lines" of variable slope joining the two components.

All terrestrial and some meteoritic samples lie on the fractionation line, but many meteoritic samples do not, suggesting the presence of an independent component. This component is most clearly characterized by the anhydrous minerals of C2 and C3 chondrites (especially CAI), which lie on a line of slope ~1 and hence must be mixtures of two components of nearly identical O^{17}/O^{18} ratio but different O^{16} content. The O^{16} -rich endmember is represented by spinel and pyroxene of $\delta O^{18} = -40\text{‰}$ and $\delta O^{17} = -42\text{‰}$. The same endmember seems to be required for ordinary chondrites and for certain anomalous inclusions [Clayton, 1981; §4(b)], and by Occam's Razor, other meteorite classes. Meteorites lying below the terrestrial line contain more of the O^{16} -rich component than does the Earth, whereas those above the line contain less. The lateral dispersion may reflect mass fractionation between solid and nebular gas. For hydrothermally altered meteorites, i.e. C1's and C2's, there seem to have been further fractionations involving liquid water (Clayton and Mayeda, 1984).

This diagram may be used to trace genetic relationships among meteorites and planets. Objects lying on the same fractionation line (e.g. Earth, Moon)

may be linked by simple mass fractionation and hence may have a common origin, either in the same body or in a homogeneous region of the nebula. Objects that do not lie on the same fractionation line contain different amounts of O^{16} and hence presumably come from different bodies. There is no simple relationship between O^{16} content and primitiveness or (presumed) heliocentric distance. C1 and C2 chondrites, both more primitive than the Earth, lie on opposite sides of the terrestrial line.

4(b) Mg, Si, Ca, Ti, Cr, Sr, Cd, Ba, Nd, Sm

The literature on the subjects of §§ 4(b) and 4(c) is vast, and cannot be fully covered in this paper. The reader may wish to consult the reviews by Arnould (1986), Shima (1986), Wood (1981), Begemann (1980), and Wasserburg et al. (1980).

Mg and Si in CAI's often show evidence of slight mass fractionation, apparently reflecting condensation, evaporation, or both (Lee and Papanastassiou, 1974; Yeh and Epstein, 1978; Molini-Velsko et al., 1983). So does Cd, a volatile trace element (Rosman et al., 1980).

All CAI's from Allende and Leoville show small ($\sim 0.8\%$) excesses of Ti^{50} (Niederer et al., 1981; Niemeyer and Lugmair, 1981, Fahey et al., 1985). Presumably a separate dust component was present, enriched in Ti^{50} (D.D. Clayton, 1981). Similar enrichments have been found for Ca^{48} (Jungck et al., 1984) and Cr^{54} (Birck and Allègre, 1984).

The remaining elements are isotopically normal in most CAI's, but show fractionation and unidentified nuclear effects in a few special inclusions, designed by the acronym "FUN" (Wasserburg et al., 1977). The FUN inclusions also have unusual oxygen isotope compositions, lying on mixing lines between a common (gas?) composition (H in Fig. 5) and several points along a mass

fractionation line AG. This line emanates from the O^{16} -rich composition A, representing the anomalous dust component in the nebula. Two models have been proposed for the FUN inclusions.

(1) Dust component A underwent mass fractionation, yielding solids B, C, D, etc. Each of these polymineralic solids then reacted with the nebular gas H, under conditions where only the most reactive minerals achieved exchange equilibrium (Blander and Fuchs, 1975). This left the minerals strung out along mixing lines BH, CH, etc. (Clayton and Mayeda, 1977; Wasserburg et al., 1977; Lee et al., 1980).

(2) Solids A, B, C, etc. are mixtures of two mass fractionated components in a presolar locale, gas to the left of A and dust to the right of G. Upon arrival in the solar system, these solids would partially equilibrate with nebular gas H, as above (D.D. Clayton, 1981).

On either model, some ad hoc reason must be found why only the highly mass fractionated solids B, C, retained nuclear anomalies; in other words, why is F coupled with UN? One possibility is that a presolar melting process greatly coarsened B, C, D, etc. compared to A, thus minimizing uptake of late, fine-grained ingredients of the solar system mix of heavy elements.

4(c) Extinct Radionuclides

Seven extinct radionuclides, with the half-lives given in parentheses, have left detectable decay products in meteorites: Ca^{41} (1.0×10^5 a), Al^{26} (7.2×10^5 a), Mn^{53} (3.7×10^6 a), Pd^{107} (6.5×10^6 a), I^{129} (1.6×10^7 a), Pu^{244} (8.1×10^7 a), and Sm^{146} (1.03×10^8 a)*. Some (Ca^{41} , Al^{26} , Mn^{53}) have been found

* For references, see Begemann (1980), Wasserburg et al. (1980), Birck and Allègre (1985), Crabb and Anders (1982), Lugmair et al. (1983), and Hutcheon et al. (1984).

only in CAI's and others only in irons (Pd^{107}), but this reflects mainly detectability, not necessarily limited distribution. These nuclides typically occur at only 10^{-2} to 10^{-5} the abundance of neighboring nuclides, and so their decay products will be detectable only if parent and daughter element have been chemically fractionated to an appreciable degree. For this reason, extinct nuclides are rarely found in matrix, which — though quite ancient — is the least fractionated part of meteorites.

The original hope that extinct radionuclides would provide a detailed relative chronology of the early solar system has not materialized. At best, one can obtain an isochron giving the ratio of an extinct nuclide to a stable isotope (e.g., $\text{Al}^{26}/\text{Al}^{27}$) in a given meteorite, referring to the time the system became closed to diffusion of the daughter (Mg^{26} in this example). But if two meteorites have different ratios, this can reflect not only differences in age but also differences in initial ratios, given the evidence for isotopic heterogeneity in the early solar system [§4(a,b)]. There is no way of separating these two variables.

In view of this ambiguity, very little can be said about the variation of extinct radionuclides with heliocentric distance. The distribution of Al^{26} can perhaps be traced by its heat effects, as the usual $\text{Al}^{26}/\text{Al}^{27}$ ratio of 5×10^{-5} would cause even ~10 km asteroids to melt. On this basis, one could argue that little or no Al^{26} got into the Earth and other terrestrial planets, as the heat effects would be staggering (Urey, 1955). However, as the accretion time of the terrestrial planets ($\sim 10^7$ a, Wetherill, 1980) is much longer than the half-life of Al^{26} (7.2×10^5 a), slow accretion would solve the problem equally well. An argument in favor of patchy distribution of Al^{26} is the difference in reflection spectra among the largest asteroids.

4 Vesta is differentiated to basalt and hence must have extensively melted, whereas 1 Ceres and 2 Pallas have relatively primitive, olivine-magnetite surfaces, implying little or no differentiation (Gaffey and McCord, 1978).

4(d) Noble Gases, C, N, and H

Several isotopically anomalous components of these elements have been found in meteorites. The noble-gas components, though present at concentrations of only $10^8 - 10^{10}$ atoms/g, were the first to be identified, and then served as tracers in the search for their (likewise exotic) carrier phases. Most of these carriers are carbonaceous, but have not yet been fully characterized or isolated in pure form. They differ from each other in noble-gas release temperature, grain size, density, combustibility, and isotopic composition. Pending further characterization, they are non-committally designated by Greek letters.

Table 2 lists the exotic noble-gas components and their carriers, as well as two exotic components of H and N that are not associated with noble gases.

Two types of Ne-E are distinguished, differing in release temperature and carrier. Ne-E(L) is virtually pure ($\geq 99\%$) Ne^{22} , and though the experimental limit for Ne-E(H) is less stringent ($\geq 83\%$), it, too, probably is pure Ne^{22} . The only plausible source of monoisotopic Ne^{22} is β^+ -decay of Na^{22} ($t_{1/2} = 2.6\text{a}$). The short half-life of Na^{22} implies that it was synthesized and trapped on an equally short time scale, i.e. under catastrophic conditions, presumably in a nova (Clayton and Hoyle, 1976; Arnould and Beelen, 1974). The high δC^{13} values certainly are consistent with a nova origin, but the expected high δN^{15} is yet to be confirmed, as

the samples analyzed were not pure enough (Lewis et al., 1983).

The association of Ne-E(H) with spinel is less certain now than it was 2 years ago. Niederer et al. (1985) have found heavy C of $\delta C^{13} > 2000\text{‰}$ in Ne-E-bearing spinel fractions. It is not clear whether this heavy carbon is present as discrete grains clinging to or enclosed in spinel or whether it is dissolved in spinel on an atomic scale.

Xenon-S shows the distinct signature of the s-process: strong predominance of even isotopes over odd ones (128, 130, 132 vs. 129, 131), rise from 128 to 132, and absence of Xe isotopes that do not lie on the s-process path (124, 126, 134, 136). As the s-process is known to take place in red giants, Xe-S must have formed in a red giant. Originally, Swart et al. (1983) inferred that Xe-S is located in C β , a type of carbon strongly enriched in C^{13} ($\delta C^{13} = +1400\text{‰}$), but recent work (Tang et al., 1986) has decoupled these two components, suggesting instead that Xe-S is located in carbon of essentially terrestrial isotopic composition (C θ , with $\delta C^{13} \approx -20\text{‰}$). Either association is consistent with astronomical observations, as s-process enrichments are observed in stars of high as well as low C^{12}/C^{13} .

Both Xe-S and C θ could have been ejected during the red giant stage (red giants lose mass at a rate of 10^{-7} to $10^{-8} M_{\odot}/a$, forming circumstellar shells of graphite grains). But it is also possible that the Xe-S was ejected only later, and was trapped by ion implantation in interstellar dust.

The nucleosynthetic origin of Xe-HL is less well understood, though a supernova, with high rates of neutron capture and photonuclear reactions, may provide the right environments for simultaneous enrichment of heavy and light isotopes (Heymann and Dziczkaniec, 1979). One interesting clue is the lack of enrichment at mass 129 relative to 128 or 130, although mass 129 — in contrast to the others — originally forms not as xenon but as a long-lived iodine isotope ($t_{1/2} = 16 \times 10^6 a$). Apparently the atoms were trapped by a chemically non-selective process, such as ion implantation (Lewis and

Anders, 1981). This could happen if the supernova ejecta,, travelling at $\sim 10^4$ km/s, overtook a shell of carbon grains ejected at the presupernova stage (D.D. Clayton, 1982). It is interesting in this connection that a significant fraction of interstellar carbon grains appears to be only 10 - 15 Å in size (Sellgren, 1984; Léger and Puget, 1984, Léger and d'Hendecourt, 1985) close to the observed 20 Å size of C6.

Deuterium enrichments of up to 5000‰ have been found in stepped combustion or pyrolysis of the organic polymer from primitive chondrites (Kolodny et al., 1981). Although such enrichments could, in principle, be achieved by equilibrium fractionations, rates are far too low at the required temperatures of ~ 100 K (Geiss and Reeves, 1981). More probably, these enrichments were caused by ion-molecule reactions (Kolodny et al., 1980; Geiss and Reeves, 1981), which are believed to be responsible for the even greater D-enrichments in interstellar molecules (Smith, 1986). The observed enrichment factor of ~ 50 relative to galactic hydrogen (Fig. 6) might represent either formation of the entire polymer by ion-molecule reactions at a rather high temperature (~ 60 K) and consequently small fractionation, or admixture of a minor amount of highly enriched material, i.e. polymerized interstellar molecules.

The latter possibility seems more likely. On acid treatment or oxidation, the D-rich material reacts preferentially, which suggests that it contributes a minor fraction of the total (McNaughton et al., 1982; Yang and Epstein, 1984). Detailed structural analyses of the polymer show that while most of it is aromatic, as discussed in §3(d), a minor part ($\sim 1\%$) is acetylenic (Hayatsu et al., 1983). The latter component yields fragments such as C_mH ($m = 3 - 10$) and C_mCN ($m = 3 - 6$), which are closely related to polycyanoacetylenes $[H(C \equiv C)_nCN]$,

a prominent class of interstellar molecules ranging up to $n = 6$.

A strikingly large enrichment of N^{15} (up to 973‰) has been found in the brecciated stony irons Bencubbin and Weatherford (Prombo and Clayton, 1985). The nature of the carrier is not known. By their oxygen isotope composition the silicates may be related to C2R and C2M chondrites, and there are suggestions that the metal clasts may be direct nebular condensates (Newsom and Drake, 1979). But otherwise these meteorites are far from primitive, having undergone severe brecciation and perhaps impact melting (Kallemeyn et al., 1978). At present, it is not possible to decide whether the nitrogen anomaly is nuclear or chemical, i.e. produced by ion-molecule reactions in a molecular cloud (Prombo and Clayton, 1985).

5. Comets

We already know that the "rock" fraction of comets is approximately chondritic, and that the "ice" fraction differs appreciably from the equilibrium composition predicted by Fig. 1, since it contains mainly CO_2 rather than CH_4 as well as formaldehyde and other "interstellar" molecules (Delsemme, 1982; Huebner et al., 1982). For further characterization of comets, we would like to know to what extent they shared the "reprocessing" of chondrites (Anders, 1986). Do comets contain chondrules, CAI, Fe^{2+} -rich silicates, hydrated minerals, magnetite, etc.? Were they ever exposed to liquid water? Have they been metamorphosed? What kinds of organic compounds do they contain? Some clues have already been obtained from meteor spectra (Wilkening, 1975) or stratospheric dust particles (Brownlee, 1986; Fraundorf and Shirck, 1979), and others will undoubtedly come from the Vega and Giotto missions. But most of the answers will require a sample return mission.

Such a mission is even more essential if we want to study the isotopic record of comets, particularly its presolar part, which ought to be clearer and more diverse than that in meteorites. Two isotopic anomalies — for Mg (Esat et al., 1979) and H (McKeegan et al., 1985) — have actually been detected in stratospheric dust particles, at least some of which are thought to be of cometary origin (Brownlee, 1986). However, these particles are so small ($<10^{-9}$ g) that only a few of the most abundant elements can be analyzed isotopically. The remainder of the cometary record — including the unique presolar part — will remain inaccessible to us until a comet sample is returned to Earth.

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REFERENCES

- Alaerts, L., Lewis, R.S., Matsuda, J. & Anders, E. 1980 *Geochim. Cosmochim. Acta* 44, 189-209.
- Allen, J.S., Nozette, S., & Wilkening, L.L. 1980 *Geochim. Cosmochim. Acta* 44, 1161-1175.
- Anders, E. 1965 *Meteoritika* 26, 17-25. English version: NASA Contractor Report, NASA CR-299.
- Anders, E. 1986 In *Comet Nucleus Sample Return*, Proc. ESA Workshop, Canterbury, England, July 15-17, in press.
- Anders, E., Hayatsu, R. & Studier, M.H. 1973 *Science* 182, 781-790.
- Anders, E. & Zadnik, M.G. 1985 *Geochim. Cosmochim. Acta* 49, 1281-1291.
- Armstrong, J.T., El Goresy, A. & Wasserburg, G.J. 1985 *Geochim. Cosmochim. Acta* 49, 1001-1022.
- Arnould, M. 1986 *Phil. Trans. Roy. Soc. [A]*, this issue.
- Arnould M. & Beelen W. 1974 *Astron. & Astrophys.* 33, 215-230.
- Ashworth, J. R. 1977 *Earth Planet. Sci. Lett.* 35, 25-34.
- Begemann, F. 1980 *Rep. Prog. Phys.* 43, 1309-1356.
- Birck, J.L. & Allegre, C.J. 1984 *Geophys. Res. Lett.* 11, 943-946.
- Birck, J.L. & Allegre, C.J. 1985 *Meteoritics* 20, 609.
- Black, D.C. & Pepin, R.O. 1969 *Earth Planet. Sci. Lett.* 6, 395-405.
- Blander, M. & Fuchs, L.H. 1975 *Geochim. Cosmochim. Acta* 39, 1605-1619.
- Brownlee, D. 1986 *Phil. Trans. Roy. Soc.[A]*, this issue.
- Cameron, A.G.W. 1973 In *Interstellar Dust and Related Topics* (ed. Greenberg, J.M. and van de Hulst, H.C.), 545-547. Dordrecht: Reidel.
- Cameron, A.G.W. 1978 In *Protostars and Planets* (ed. T. Gehrels), 453-487, Tucson: University of Arizona Press.
- Carr, R.H., Wright, I.P., Pillinger, C.T., Lewis, R.S., & Anders, E. 1983 *Meteoritics* 18, 277.

- Clayton, D.D. 1975 *Astrophys. J.* **199**, 765-769.
- Clayton, D.D. 1981 *Astrophys. J.* **251**, 374-386.
- Clayton, D.D. 1982 *Quart. J. Roy. Astron. Soc.* **23**, 174-212.
- Clayton, D.D. & Hoyle, F. 1976 *Astrophys. J.* **203**, 490-496.
- Clayton, R.N. 1981 *Phil. Trans. Roy. Soc. London [A]* **303**, 339-349.
- Clayton, R.N., Grossman, L. & Mayeda, T.K. 1973 *Science* **182**, 485-488.
- Clayton, R.N., MacPherson, G.J., Hutcheon, A.M., Davis, L., Grossman, L., Mayeda, T.K., Molini-Velsko, C., Allen, J.M., & El Goresy, A. 1984 *Geochim. Cosmochim. Acta* **48**, 535-548.
- Clayton, R.N. & Mayeda, T.K. 1977 *Geophys. Res. Lett.* **4**, 295-298.
- Clayton, R.N. & Mayeda, T.K. 1984 *Earth Planet. Sci. Lett.* **67**, 151-161.
- Crabb, J., Lewis, R.S., & Anders E. 1982 *Geochim. Cosmochim. Acta* **46**, 2511-2526.
- Delsemme, A.H. 1982 In *Comets* (ed. Wilkening, L.L.), pp. 85-130. Tucson: University of Arizona Press.
- DuFresne, E.R. & Anders, E. 1962 *Geochim. Cosmochim. Acta* **26**, 1085-1114.
- El Goresy, A., Nagel, K. & Ramdohr, P. 1978 *Proc. Lunar Sci. Conf.* **9**, 1279-1303.
- Esat, M., Brownlee, D.E., Papanastassiou, D.A., & Wasserburg, G.J. 1979 *Science* **206**, 190-197.
- Fahey, A., Goswami, J.N., McKeegan, K.D., & Zinner, E. 1985 *Astrophys. J.* **296**, L17-L20.
- Fraundorf, P. & Shirck, J. 1979 *Proc. Lunar Planet. Sci. Conf.* **10**, 951-976.
- Gaffey, M.J. & McCord, T.B. 1978 *Space Science Reviews* **21**, 555-628.
- Galwey, A. 1972 *Geochim. Cosmochim. Acta* **36**, 1115-1130.
- Geiss, J. & Reeves, H. 1981 *Astron. Astrophys.* **93**, 189-199.
- Gooding, J.L. & Keil, K. 1980 *Meteoritics* **16**, 17-43.

- Grossman, J.N. Rubin, A.E., Rambaldi, E.R., Rajan, R.S. & Wasson, J.T. 1985 *Geochim. Cosmochim. Acta* 49, 1781-1795.
- Grossman, J.N. & Wasson, J.T. 1983 *Geochim. Cosmochim. Acta* 47, 759-771.
- Grossman, J.N. & Wasson, J.T. 1985 *Geochim. Cosmochim. Acta* 49, 925-939.
- Grossman, L. 1972 *Geochim. Cosmochim. Acta* 36, 597-619.
- Grossman, L. 1980 *Ann. Review Earth Planet. Sci.* 8, 559-608.
- Grossman, L. & Ganapathy, R. 1976 *Geochim. Cosmochim. Acta* 40, 967-977.
- Grossman, L. & Ganapathy, R., & Davis, A.M. 1977 *Geochim. Cosmochim. Acta* 41, 1647-1664.
- Grossman, L. & Larimer, J.W. 1974 *Rev. Geophys. Space Phys.* 12, 71-101.
- Hayashi, C., Nakazawa, K. & Nakagawa, Y. 1985 In *Protostars and Planets II* (ed. Black, D.C. & Matthews, M.S.), pp. 1100-1153.
- Hayatsu, R. & Anders, E. 1981 *Topics of Current Chemistry* 99, 1-34.
- Hayatsu, R., Matsuoka S., Scott, R.G., Studier, M.H. & Anders E. 1977 *Geochim. Cosmochim. Acta* 41, 1325-1339.
- Hayatsu, R., Scott, R.G., & Winans, R.E. 1983 *Meteoritics* 18, 310.
- Heymann, D. & Dziczkaniec, M. 1979 *Proc. Lunar Planet Sci. Conf.* 10, 1943-1959.
- Huebner, W.F., Giguere, P.T., & Slattery, W.L. 1982 In *Comets* (ed. Wilkening, L.L.), pp. 496-515. Tucson: University of Arizona Press.
- Huss, G.R. 1986 *J. Geophys. Res.*, in press.
- Huss, G.R., Keil, K., & Taylor, G.J. 1981 *Geochim. Cosmochim. Acta* 45, 33-51.
- Hutcheon, I.D., Armstrong, J.T., & Wasserburg, G.J. 1984 *Lunar Planet. Sci.* 15, 387-388.
- Hutchison, R., Bevan, A.W.R., Agrell, S.O., & Ashworth, J.R. 1979 *Nature* 280, 116-119.
- Jungck, M.H.A. & Eberhardt, P. 1979 *Meteoritics* 14, 439-440.

- Jungck, M.H.A., Shimamura, T., & Lugmair, G.W. 1984 *Geochim. Cosmochim. Acta* 48, 2651-2658.
- Kallemeyn, G.W., Boynton, W.V., Willis, J. & Wasson, J.T. 1978 *Geochim. Cosmochim. Acta* 42, 507-515.
- Kolodny, Y., Kerridge, J.F., & Kaplan, I.R. 1980 *Earth Planet. Sci. Lett.* 46, 149-158.
- Kornacki, A.S. & Wood, J.A. 1984 *J. Geophys. Res.* 89, B573-B587.
- Kurat, G. 1970 *Earth Planet. Sci. Lett.* 9, 225-231.
- Larimer, J.W. 1967 *Geochim. Cosmochim. Acta* 31, 1215-1238.
- Larimer, J.W. 1975 *Geochim. Cosmochim. Acta* 39, 389-392.
- Larimer, J.W. 1978 In *The Origin of the Solar System* (ed. Dermott, S.F.), pp. 347-393. New York: Wiley.
- Larimer, J.W. & Anders, E. 1967 *Geochim. Cosmochim. Acta* 31, 1239-1270.
- Larimer, J.W. & Anders, E. 1970 *Geochim. Cosmochim. Acta* 34, 367-388.
- Larimer, J.W. & Bartholomay, H.A. 1983 *Lunar Planet. Sci.* 14, 423.
- Larimer, J.W. & Bartholomay, M. 1979 *Geochim. Cosmochim. Acta* 43, 1455-1466.
- Lattimer, J.M. Schramm, D.N., & Grossman, L. 1978 *Astrophys. J.* 219, 230-249.
- Lee, T., Mayeda, T.K., & Clayton, R.N. 1980 *Geophys. Res. Lett.* 7, 493-496.
- Lee, T. & Papanastassiou, D.A. 1974 *Geophys. Res. Lett.* 1, 225-228.
- Léger, A. & d'Hendecourt, L. 1985 *Astron. Astrophys.* 146, 81-85.
- Léger, A. & Puget, J.L. 1984 *Astron. Astrophys.* 137, L5-L8.
- Lewis, J.S. 1972 *Icarus* 16, 241-252.
- Lewis, R.S. & Anders, E. 1981 *Astrophys. J.* 247, 1122-1124.
- Lewis, R.S., Anders, E., Wright, I.P., Norris, S.J., & Pillinger, C.T. 1983 *Nature* 305, 767-771.
- Lewis, R.S., Srinivasan, B. & Anders, E. 1975 *Science* 190, 1251-1262.

- Lugmair, G.W., Shimamura, T., Lewis, R.S. & Anders, E. 1983 *Science* 222, 1015-1018.
- Marvin, U.B., Wood, J.A., & Dickey, J.S., Jr. 1970 *Earth Planet. Sci. Lett.* 7, 346-350.
- McKeegan, K.D., Walker, R., & Zinner, E. 1985 *Geochim. Cosmochim. Acta* 49, 1971-1987.
- McNaughton, N.J., Fallick, A.E., & Pillinger, G.T. 1982 *J. Geophys. Res.* 87, A297-A302.
- Molini-Velsko, C., Mayeda, T.K., & Clayton, R. N. 1983 *Lunar Planet. Sci.* 14, 509-510.
- Morgan, J.W. & Anders, E. 1980 *Proc. Natl. Acad. Sci. U.S.* 77, 6973-6977.
- Nagahara, H. 1981 *Nature* 292, 135-136.
- Nagahara, H. 1983 *Proc. 8th Symp. Antarctic Meteorites*, 61-84.
- Nagahara, H. 1984 *Geochim. Cosmochim. Acta* 48, 2581-2595.
- Newsom, H.E. & Drake, M. 1979 *Geochim. Cosmochim. Acta* 43, 689-707.
- Niederer, F.R., Eberhardt, P., Geiss, J., & Lewis, R.S. 1985 *Meteoritics* 20, 716-717.
- Niederer, F.R. & Papanastassiou, D.A. 1984 *Geochim. Cosmochim. Acta* 48, 1279-1293.
- Niederer, F.R., Papanastassiou, D.A., & Wasserburg, G.J. 1981 *Geochim. Cosmochim. Acta* 45, 1017-1031.
- Niemeyer, S. & Lugmair, G.W. 1981 *Earth Planet. Sci. Lett.* 53, 211-225.
- Osborn, T.W., Warren, R.G., Smith, R.H., Wakita, H., Zellmer, D.L., & Schmitt, R.A. 1974 *Geochim. Cosmochim. Acta* 38, 1359-1378.
- Ott, U., Yang, J. & Epstein, S. 1985 *Meteoritics* 20, 722-723.
- Palme, H. & Wlotzka, F. 1976 *Earth Planet. Sci. Lett.* 33, 45-60.
- Planner, H.N. & Keil, K. 1982 *Geochim. Cosmochim. Acta* 46, 317-330.
- Prombo, C.A. & Clayton, R.N. 1985 *Science* 230, 935-937.

- Rambaldi, E.R., Housley, R.M., & Rajan, R.S. 1984 *Nature* 311, 138-140.
- Rambaldi, E.R., Rajan, R.S., Wang, D. & Housley, R.M. 1983 *Earth Planet. Sci. Lett.* 66, 11-24.
- Rambaldi, E.R. & Wasson, J.T. 1984 *Geochim. Cosmochim. Acta* 48, 1885-1897.
- Reynolds, J.H. & Turner, G. 1964 *J. Geophys. Res.* 69, 3263-3281.
- Robert, F., Merlivat, L. & Javoy, M. 1979 *Nature* 282, 785-789.
- Rosman, K.J.R., deLaeter, J.R., & Gorton, M.P. 1980 *Earth Planet. Sci. Lett.* 48, 166-170.
- Schmitt, R.A., Smith, R.H., & Goles, G.G., 1965 *J. Geophys. Res.* 70, 2419-2444.
- Sears, D.W., Grossman, J.N., Melcher, C.L., Ross, L.M., & Mills, A.A. 1980 *Nature* 287, 791-795.
- Sellgren, K. 1984 *Astrophys. J.* 277, 623-633.
- Shima, M. 1986 *Geochim. Cosmochim. Acta* 50, 577-584.
- Smith, D. 1986 *Phil. Trans. Roy. Soc. [A]*, this issue.
- Srinivasan, B. & Anders, E. 1978 *Science* 201, 51-56.
- Studier, M.H., Hayatsu, R., & Anders, E. 1968 *Geochim. Cosmochim. Acta* 32, 151-174.
- Swart, P.K., Grady, M.M., Pillinger, C.T., Lewis, R.S., & Anders, E. 1983 *Science* 220, 406-410.
- Tang, M., Lewis, R.S., Anders, E., Grady, M.M., Wright, I.P., & Pillinger, C.T. 1986 Paper presented at Meteoritical Soc. Mtg., New York, NY, September 22-25.
- Tsuchiyama, A., Nagahara, H., & Kushiro, I. 1980 *Earth Planet. Sci. Lett.* 48, 155-165.
- Tsuchiyama, A., Nagahara, H., & Kushiro, I. 1981 *Geochim. Cosmochim. Acta* 45, 1357-1367.
- Urey, H.C. 1955 *Proc. Nat. Acad. Sci. U.S.* 41, 127-144.

- Van Schmus, W.R. & Wood, J.A. 1967 *Geochim. Cosmochim. Acta* 31, 747-765.
- Wark, D.A. & Lovering, J.F. 1982 *Geochim. Cosmochim. Acta* 46, 2581-2594.
- Wasserburg, G.J., Lee, T., & Papanastassiou, D.A. 1977 *Geophys. Res. Lett.* 4, 299-302 (1977).
- Wasserburg, G.J., Papanastassiou, D.A., & Lee, T. 1980 In *Early Solar System Processes and the Present Solar System*, pp. 144-191. Soc. Italiana di Fisica.
- Wetherill, G.W. 1980 *Ann. Rev. Astron. Astrophys.* 18, 77-113.
- Wilkening, L.L. 1975 *Nature* 258, 689-690.
- Wood, J.A. 1958 *Smithsonian Astrophys. Obs. Tech. Report* #10.
- Wood, J.A. 1962 *Geochim. Cosmochim. Acta* 26, 739-749.
- Wood, J.A. 1973 *Icarus* 2, 152-180.
- Wood, J.A. 1981 *Earth Planet. Sci. Lett.* 56, 32-44.
- Wood, J.A. 1984 *Earth Planet. Sci. Lett.* 70, 11-26.
- Wood, J.A. 1985 *Meteoritics* 20, 787-788.
- Wood, J.A. 1986 *Lunar Planet. Sci.* 17, 956-957.
- Yang, J. & Epstein, S. 1984 *Nature* 311, 544-547.
- Yeh, H.W. & Epstein, S. 1978 *Lunar Planet. Sci.* 9, 1289-1291.

TABLE 1. CARBON IN CHONDRITES

Type	Temperature, K		CARBON				H	N
	Cond.	Metam.	Total ppm	Organic %	C [*] %	CO ₃ [≡] %		
C1	360	--	36000	95	~1	~4	7900	2800
C2	400	--	23000	95	~1	~2	8900	1500
C3V	420	600	6700	≤20	>80	-	500	61
H4	450	900	1300	<10	>90	-	5	47
H5	470	1000	1100	<10	>90	-	4	43
H6	490	1200	1060	<10	>90	-	4	50

* Elemental C or solid solution of C in γ-Fe

TABLE 2. EXOTIC COMPONENTS OF NOBLE GASES, C, H, AND N

Component	Carrier	Enriched in	Release T °C	$\delta^{13}\text{C}$		$\delta^{15}\text{N}$		Abundance ppm	Size μm	Probable Source	Reference
				‰	‰	‰	‰				
Ne-E(H)	Spinel	Ne ²²	1200	>2000			1700		1-10	Nova?	aei
Ne-E(L)	Ca	Ne ²²	600	340		>0	5		1-10	Nova	aceo
	CB			1400		>0	2		0.1-3	Red Giant?	op
Xe-HL	Cδ	Xe ^{124,136}	1000	-38	-330		400		0.002	Supernova	ghlo
Xe-S	Cθ	Xe ¹³⁰	1400	-20?						Red Giant?	ajng
D	Polymer*	D	400	?	?		300		?	Molecular Cloud	dfmp
Heavy N	?				973		>60			Nova?	j

*Minor (acetylenic?) component of meteoritic polymer.

- a. Alaerts et al. (1980) i. Niederer et al. (1985)
- b. Black & Pepin (1969) j. Ott et al. (1985)
- c. Carr et al. (1983) k. Prombo & Clayton (1985)
- d. Hayatsu et al. (1983) l. Reynolds & Turner (1964)
- e. Jungck & Eberhardt (1979) m. Robert et al. (1979)
- f. Kolodny et al. (1980) n. Srinivasan & Anders (1978)
- g. Lewis et al. (1975) o. Swart et al. (1983)
- h. Lewis et al. (1983) p. Yang & Epstein (1984)
- q. Tang et al. (1986)

Figure Captions

Fig. 1. Equilibrium condensation of a solar gas explains the composition of planets, at least to first order. Inner planets consist of rock, comets and satellites of outer planets consist of rock and ice, and the outer planets consist of all 3.

Fig. 2. "Rock" portion of the diagram in Fig. 1, on linear scales (after Grossman and Larimer, 1974). Below ~700 K, metal begins to react with H_2O and H_2S in the nebula, yielding FeO (which enters silicates) and FeS. The mineral assemblage at 400 - 500 K resembles that of ordinary chondrites. Below 400 K, many major changes take place, yielding a mineral assemblage resembling that of carbonaceous chondrites.

Fig. 3. Chemical state of carbon in the solar nebula (after Hayatsu and Anders, 1981). Each solid line gives the temperatures at which 50% of the carbon should have reacted according to the equilibrium shown. In each field, the principal stable products are shown in roman type and metastable products, in italics; those that do not form for kinetic reasons are enclosed in brackets. This diagram applies to isobaric cooling as well as to isothermal compression.

CO survives down to 400 K, owing to the lack of good catalysts. At this temperature, clays form from anhydrous silicates, catalyzing hydrogenation of CO to complex organic compounds (Fischer-Tropsch reaction).

The chemical state of carbon in C-chondrites agrees with that predicted from their formation conditions (indicated by boxes), as inferred from various cosmothermometers and -barometers (Larimer, 1978). C3 chondrites contain mainly elemental carbon, whereas C1 and C2 chondrites contain mainly organic carbon (Table 1).

Fig. 4. Three-isotope plot shows variation of meteoritic O^{18}/O^{16} and O^{17}/O^{16} relative to standard mean ocean water (=SMOW). Inset gives total observed range for individual minerals; main graph shows bulk meteorites or matrix. All terrestrial samples lie on a mass-fractionation line of slope 1/2. The anhydrous minerals of C2-C4 chondrites, on the other hand, lie on a mixing line of slope ~1, corresponding to constant O^{17}/O^{18} but variable O^{16} . The latter presumably comes from an O^{16} -rich component, typified by Allende spinel (star in inset). All meteorites fall off the terrestrial line to varying degrees, indicating that their O^{16} contents differ from that of the Earth and hence preclude a genetic relationship.

Fig. 5. A few exceptional inclusions fall off the normal mixing line, and show mass fractionation and unidentified nuclear effects (=FUN). Bulk samples and separated minerals of each inclusion form linear arrays converging on a common point H. Presumably these inclusions first formed by mass fractionation of dust (composition A) along the line AG, and later partially exchanged with nebular gas H. Spinel, with the slowest exchange rate, remained on the original fractionation line, whereas more reactive minerals (melilite, pyroxene) moved toward H.

Fig. 6. (Adapted from Geiss and Reeves, 1981) Organic polymer fractions from C-chondrites are strongly enriched in deuterium, up to 32x relative to galactic hydrogen. Equilibrium isotope fractionations can produce such enrichments, but only at temperatures of ~100 K, where rates are prohibitively slow (e.g. the half-time of the Fischer-Tropsch reaction at 110 K is ~ 10^{30} years!) More probably, the fractionation is due to ion-molecule reactions, which are thought to be responsible for the large (up to 10^5 -fold) D-enrichments in interstellar molecules. Only a minor, reactive part of the meteoritic

polymer seems to be enriched in D, and probably represents relict interstellar molecules. The major part appears to be of local origin, as outlined in Fig. 3.

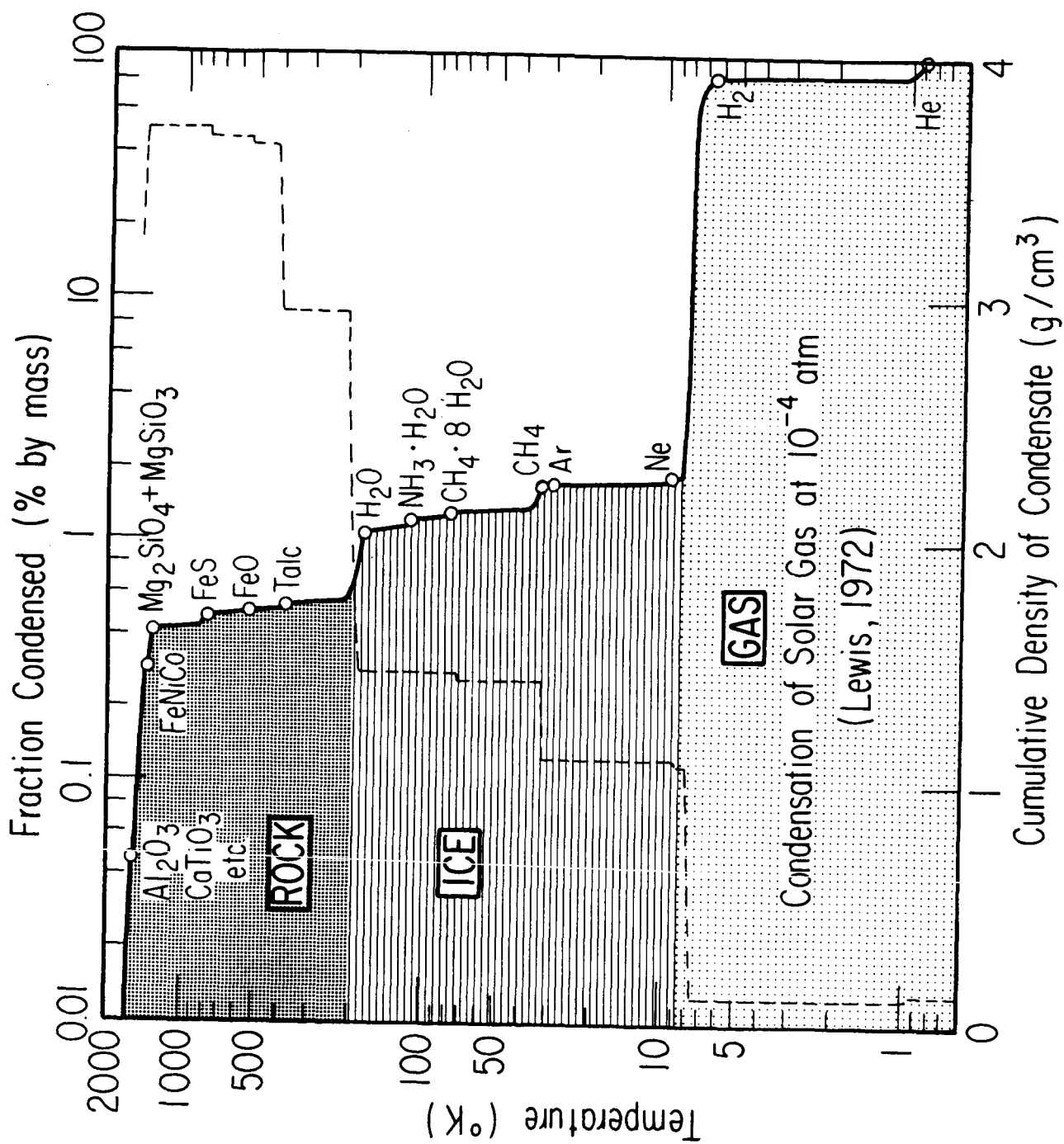


Fig.1

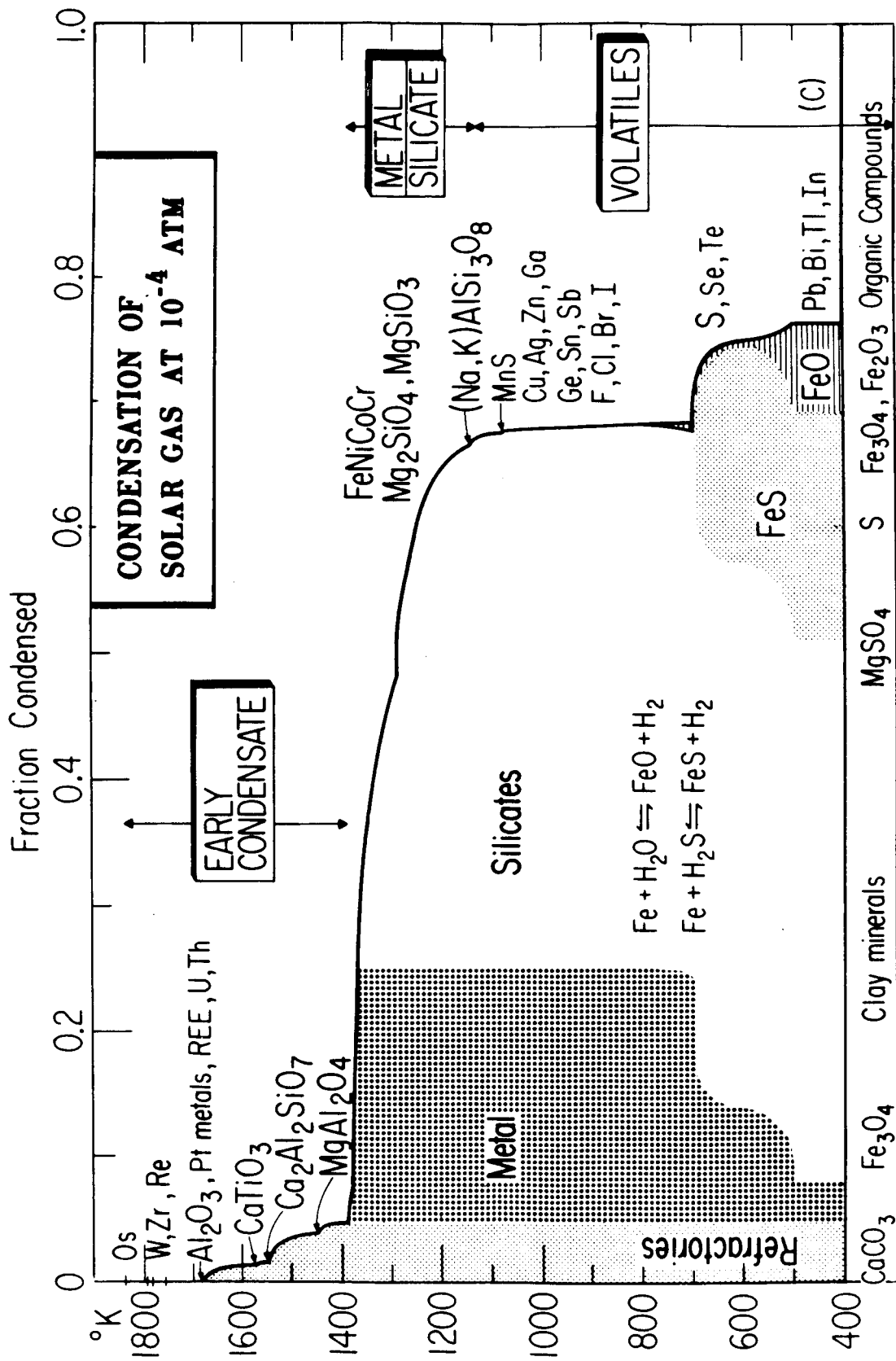


Fig.2

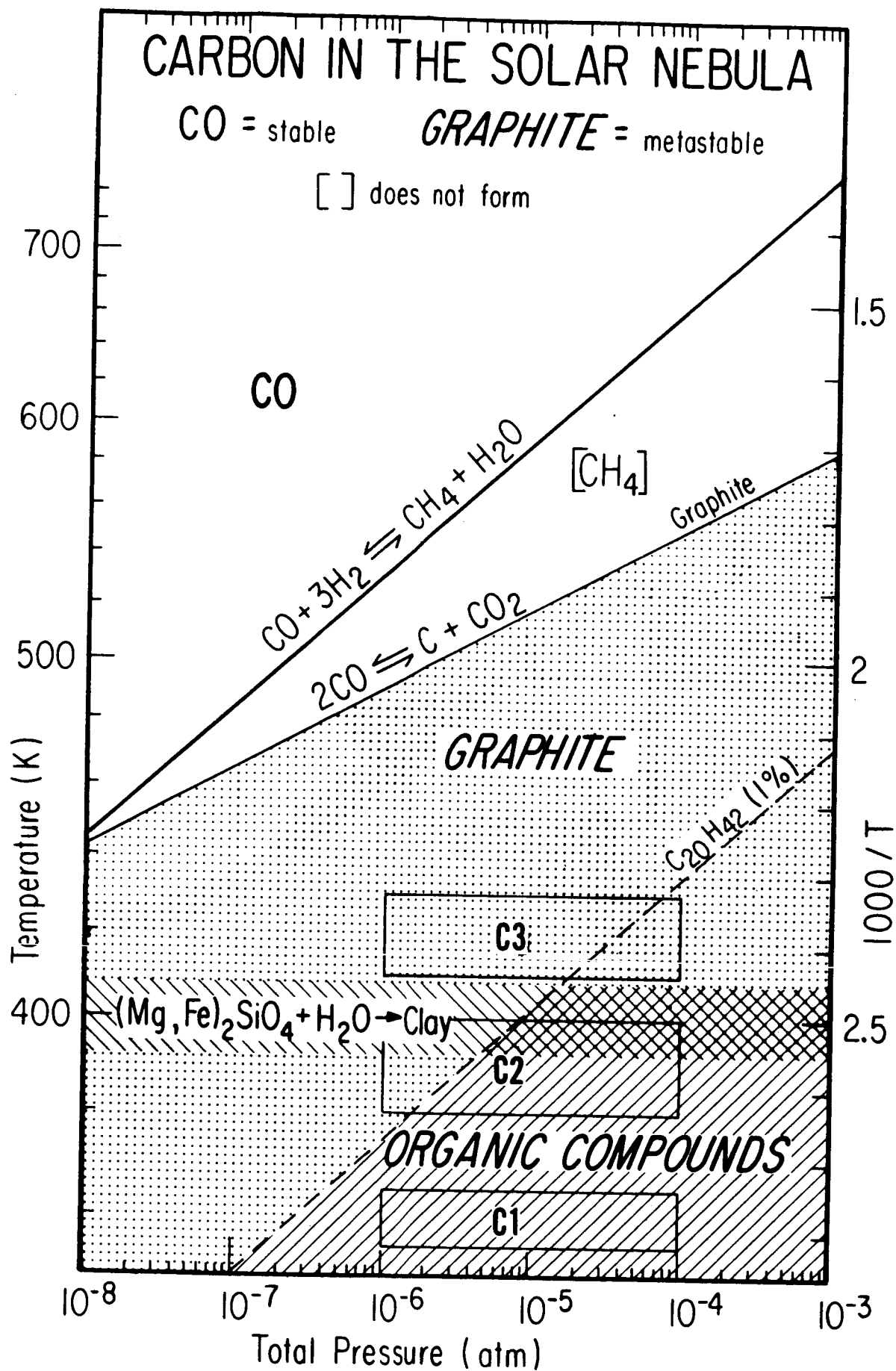


Fig.3

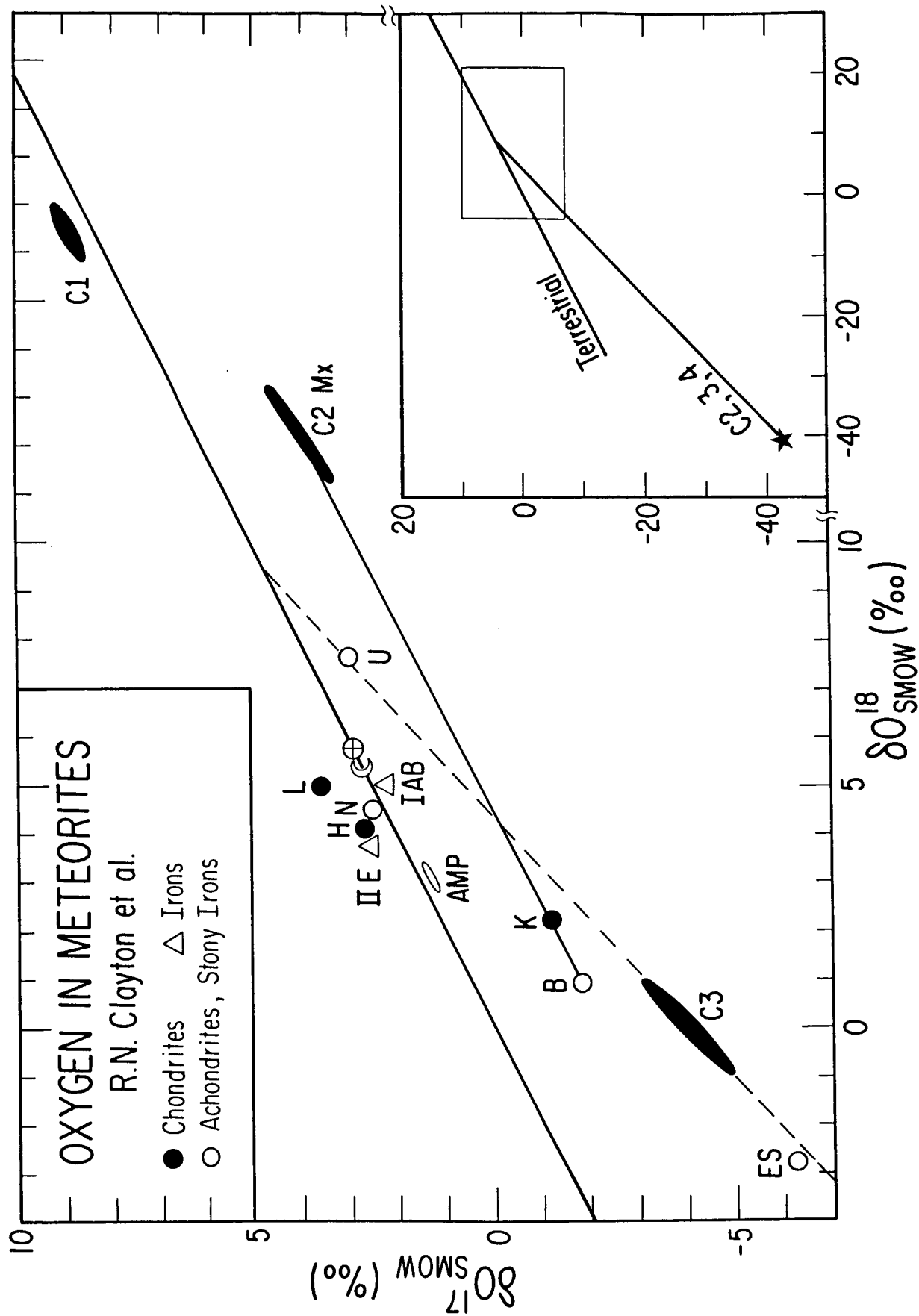


Fig. 4

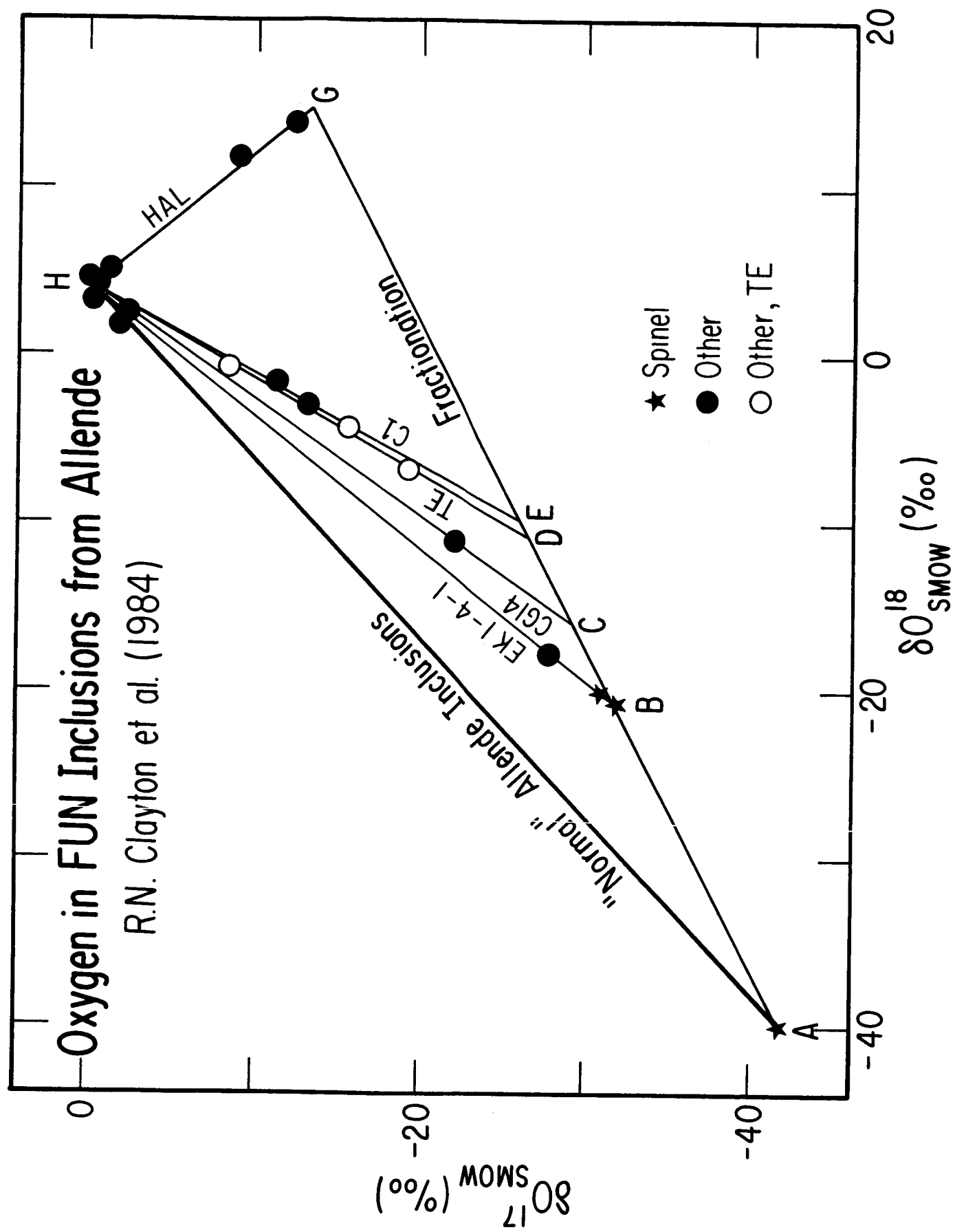


Fig.5

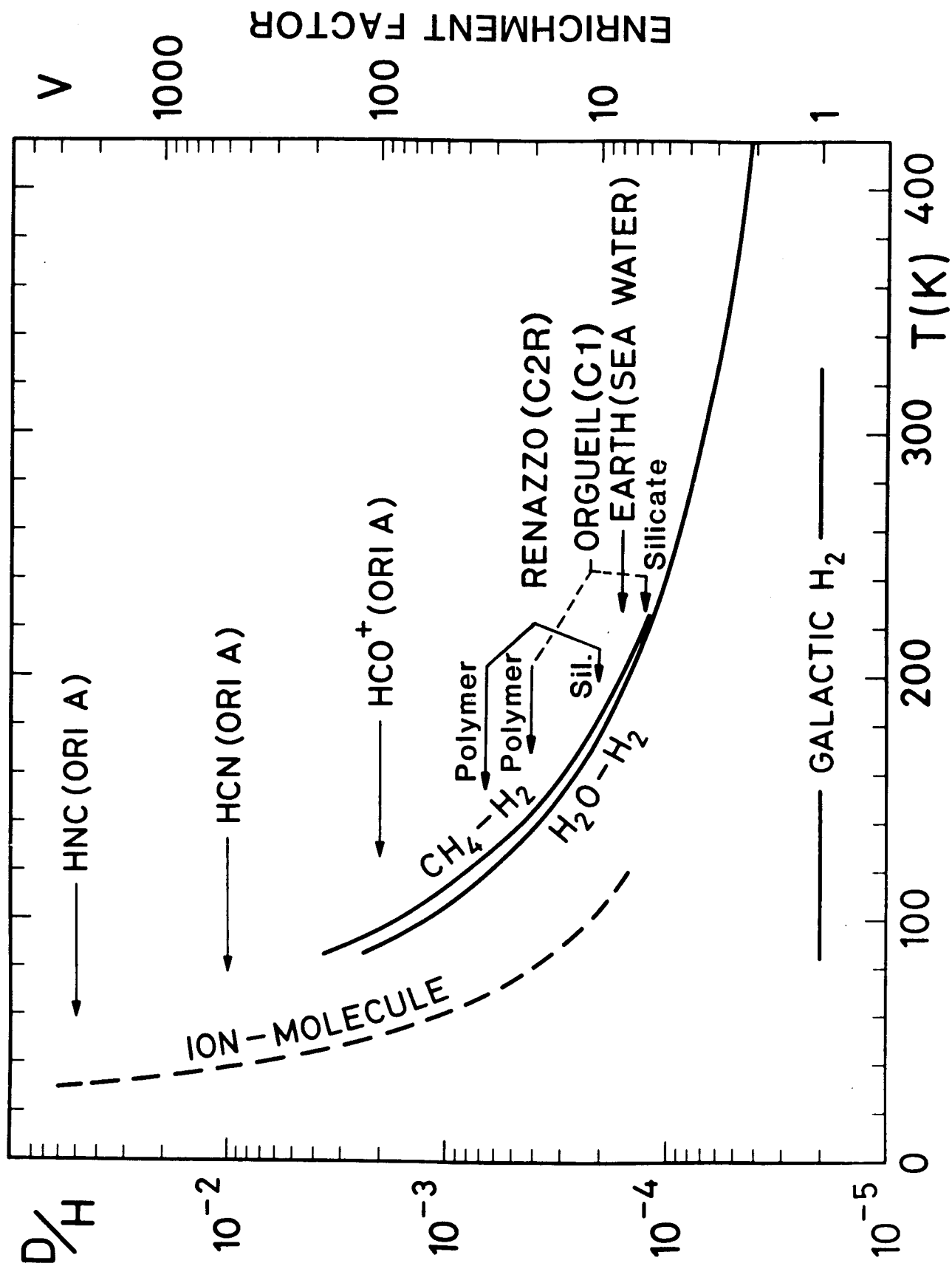


Fig.6